THE ROLE OF FLUVIAL AND GLACIAL EROSION IN LANDSCAPE EVOLUTION: THE BEN OHAU RANGE, NEW ZEALAND

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ABSTRACT

A morphometric comparison of valleys has been made for the Ben Ohau Range in the central Southern Alps of New Zealand. The range is undergoing rapid tectonic transport and uplift. The humid north of the range is a glacial trough-and-arête landscape, with a temperate glacial climate. The dry south has rounded divides and plateau remnants dissected by fluvial valleys. Assuming that space–time substitution allows today's spatial valley-form transition to represent evolutionary stages in valley development, the tectonic history allows time constraints to be placed on the rate of transition to an alpine glacial landscape. Morphometric change has been quantified using hypsometric curves, and distance–elevation plots of cirque and valley-floor altitudes. Ancestral fluvial valleys have less concave long profiles but are stepped at altitude owing to the presence of high-level cirques and remnant plateau surfaces, and possess a low proportion of land area at low elevation. Increasing glacial influence is manifest as smoother, more deeply concave long profiles and U-shaped cross-profiles associated with a higher proportion of the land area at lower elevation. The full morphological transition has involved up to 2.4 km of vertical denudation over the 4 Ma lifetime of the mountain range, of which 80 per cent would have occurred by preglacial fluvial erosion. Combining the trajectory of tectonic transport with reconstructed glaciation limits and climatic history, it is indicated that about 200 ka of temperate glacial erosion produces recognizable trough-and-arête topography. Mean and modal relief increase where glacial activity is confined to cirques, but decrease when trough incision by ice becomes established as a dominant process in the landscape. © 1997 by John Wiley & Sons, Ltd.

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INTRODUCTION

Geomorphological evidence, such as high summit elevations and deep incision, conventionally used to infer recent uplift of mountain ranges, may be ambiguous because the same effects could be caused by increased alpine glaciation following climate change (Molnar and England, 1990). The transformation of valleys from a fluvial (V-shaped) cross-profile to a fully glacial (U-shaped) cross-profile has been modelled by Harbor *et al.* (1988). They have demonstrated that a steady state, parabolic valley shape may be reached within *c.* 100 ka from onset of glaciation, given that glacial erosion rates are *c.* 10^{-3}ma^{-1}. Such glacial trough formation affects the area–altitude distribution, and is therefore pertinent to Molnar and England's (1990) consideration of late Cenozoic mountain range development. However, Koons (1995) notes that the understanding of the evolution of mountain belts is seriously hindered by a lack of field studies of glacial and fluvial systems at appropriate scales.

We present a field test of the effect of the onset of glaciation on relief development in a rapidly uplifting mountain range in New Zealand. First, the sequence of topographic transitions caused by a change in dominant process regime, from fluvial to glacial erosion, is described. Then, the time-scale of change is evaluated and compared with Harbor *et al.*'s (1988) estimate. Finally, observations are related to the hypothesized relief development implicit in existing tectonic–isostatic models of alpine uplift.

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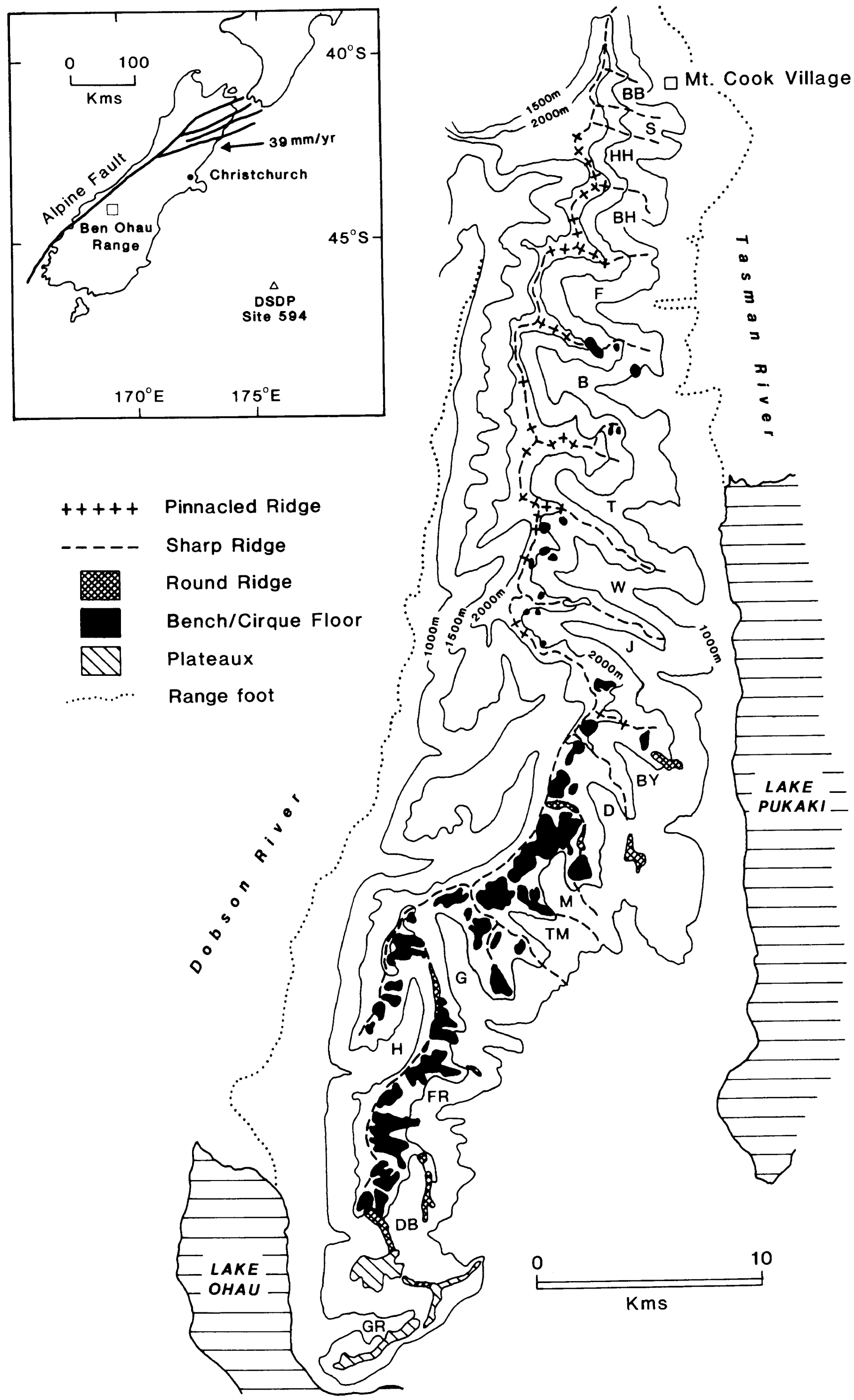

Figure 1. Location and simplified geomorphology of the Ben Ohau Range. Identification letters for each of the study valleys are as follows: GR, Gretas Stream; DB, Darts Bush Stream; FR, Fraser Stream; H, Harris Stream; G, Gladstone Stream; TM, Top McMillan Stream; M, Mackenzie Stream; D, Duncan Stream; BY, Boundary Stream; J, Jacks Stream; W, Whale Stream; T, Twin Stream; B, Bush Stream; F, Freds Stream; BH, Birch Hill Stream; HH, Hoophorn Stream; S, Sawyer Stream; BB, Black Birch Stream.

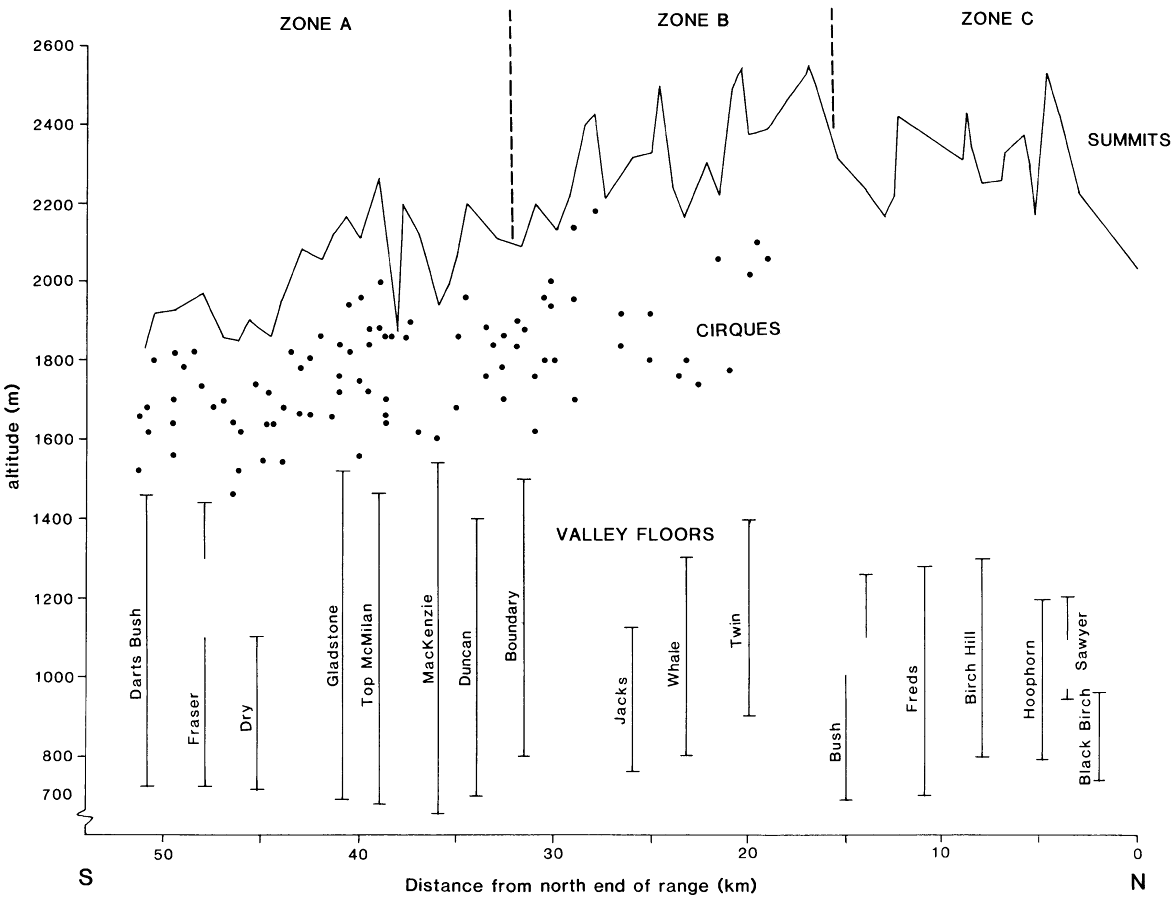

Figure 2. Altitude–distance profile of the range from south (left) to north (right), showing summit elevation, bench/cirque floor altitude, and valley floor altitude. The upper limits of valley floors are identified by the location of the change-of-slope between valley-floor and valley-head segments of long profiles constructed for each valley from 1:50 000 scale topographic maps (not shown). Such changes-of-slope are more pronounced with increasing glacial influence.

STUDY AREA

The Ben Ohau Range (Figure 1) forms a 60 km long southern offshoot of the Main Divide of the Southern Alps. The location, relief, geological history and climate gradient together provide an opportunity to study the relief development of an alpine range under more controlled conditions than are normally encountered in the field. The range is a fault-bounded uplifting block of Triassic quartzo-feldspathic greywackes and argillaceous metasediments, Upper Palaeozoic schists, and localized volcanics (Spörli and Lillie, 1974). As part of the Pacific Plate, the range is being carried to the west-southwest towards the junction with the Australian–Indian Plate along the Alpine Fault. Ridgeline elevation rises towards the Main Divide, from *c.* 1900 m in the south to *c.* 2400 m in the north (Figure 2). Precipitation increases from less than 600 mm a^{-1} in the south to over 5000 mm a^{-1} in the north (Fitzharris, 1988). Valleys draining into the Tasman Valley are the subject of this study (Table I), being broadly comparable in dimension, aspect and geology.

VALLEY MORPHOMETRY

Area–altitude (hypsometric) distributions of 18 valleys, including all the eastward drainage of the range (Table I; Figure 3A) were derived from digitizing contours at 200 m intervals from 1:50 000 scale topographic

Table I. Characteristics of catchments in the Ben Ohau Range arranged from north to south.

Name	Area	Elevation (m a.s.l.)			
	(km^2)	Max.	Min.	Mean	Modal*
Black Birch	4·65	*c.* 2200	880	1380	1100
Sawyer	3·64	2235	920	1420	1300
Hoophorn	8·27	2557	780	1460	1300
Birch Hill	9·91	2444	760	1500	1100
Freds	15·52	2444	680	1590	1900
Bush	19·36	2557	660	1670	1900
Twin	15·51	2551	900	1740	2100
Whale	21·34	2500	780	1600	1700
Jacks	15·06	2500	900	1750	1900
Boundary	14·07	2218	820	1490	1300
Duncan	16·53	2200	840	1550	1700
Mackenzie	14·07	2200	900	1560	1700
Top McMillan	8·64	2196	920	1590	1700
Gladstone	20·31	2263	800	1560	1700
Harris	19·68	2263	940	1570	1700
Fraser	13·56	1974	760	1390	1700
Darts Bush	10·80	*c.* 1940	820	1450	1500
Gretas	13·63	1690	620	1190	1100

* Modal elevation is the mid-point of the modal 200 m altitudinal band

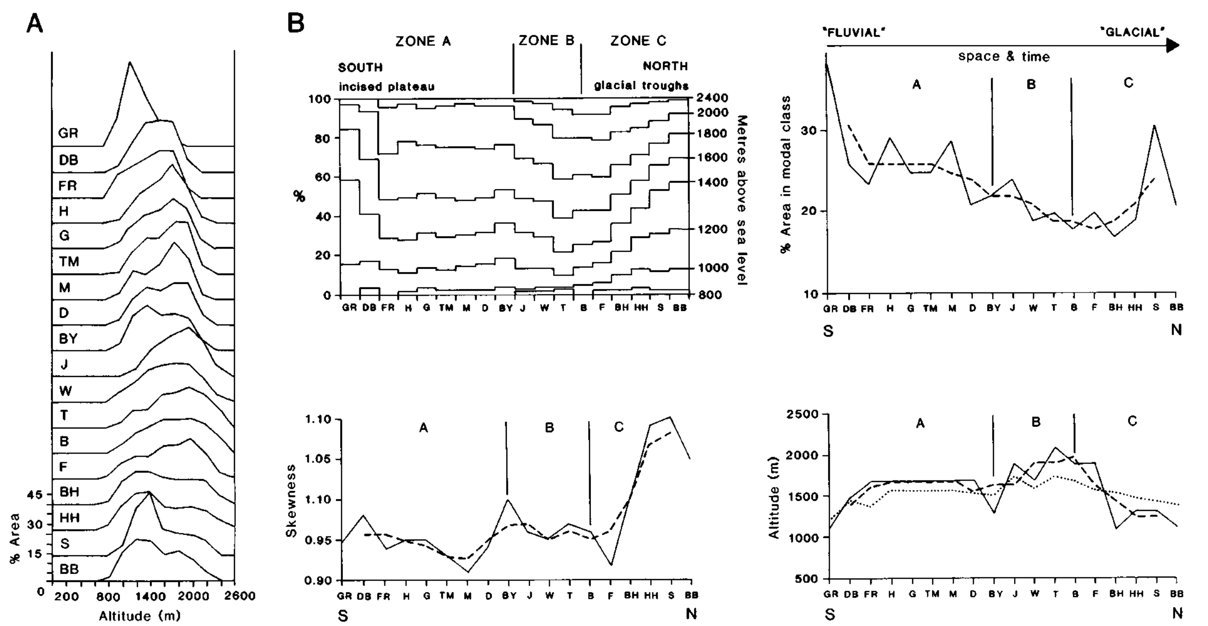

Figure 3. Morphometric characteristics of 18 valleys in the Ben Ohau Range. Letter codes are as for Figure 1. (A) Hypsometric histograms showing the distribution of area with altitude for individual valleys, arranged from south (top) to north (bottom). Percentage area is plotted at the mid-point of each 200 m altitude class (*i.e.* at 100, 300, 500 m *etc.*). (B) Derived morphometric parameters. Clockwise from top left: proportion of area with altitude for each valley from south (left) to north (right); variation in the dominance of the modal class of the hypsometric distribution; variation in the mean altitude (dotted line) and modal altitude class (solid line); and variation in skewness (Krumbein and Pettijohn method). In all cases, dashed lines indicate three-point running means (except for mean altitude, for which running mean is not shown).

maps. A hypsometric approach has the advantage of including the whole catchment in the calculation, effectively providing an integration of an infinite number of cross-profiles from each valley, and providing information in a standardized form for use in topographic and tectonic models.

Table II. Subdivision of the Ben Ohau Range into three zones on the basis of relief and landform characteristics.

Zone	Relief				Skewness	Landform
	Max.	Mean.	Mode	Min.		
A	Increasing	Increasing	Constant at high elevation; lower % of land area to north.	Constant	Positive	Plateau remnants in south transitional to bevelled sloping benches *c.* 100 m below ridgelines in north. Rounded divides, V-shaped valleys with little or no valley floor development. Dominance of fluvial erosional processes.
B	Increasing	Increasing	Increasing at high elevation; lower % land area to north.	Constant	Less positive	Sharp ridgelines, benches transitional to well-developed cirques of increasing depth and floor altitude to north. Lower valleys U-shaped with low radius of curvature. Significant and identifiable glacial influence on landforms.
C	Constant (increasing)	Decreasing	Decreasing at low elevation; higher % land area to north.	Constant	Negative	Major trough development, with increasing cross-sectional radius to north. Pinnacled ridgelines. Distinct change of slope from valley floor to headwall. Cirques broad, shallow and steep-floored or absent. Steep valley slopes but less talus than Zone B. Greater relative relief; alpine glacial landscape.

Results

Northward changes in valley morphology are not gradual and continuous, but are marked by an abrupt change in area–altitude distribution, allowing division of the range into three zones (Table II; Figure 3B). These morphometric changes can be related to landform (Figures 1 and 2).

Zone A is a landscape of fluvial incision, with rounded divides, plateau remnants, and erosional benches cut up to 150 m below the ridgelines, transitional to shallow cirques to the north. Zones A and B are distinguished on morphological grounds, namely the replacement of rounded divides and plateau remnants by sharp ridgelines, rather than by a morphometric transition. Thus, relict landsurfaces exist in Zone A but are absent from B. Throughout Zones A and B, summit and cirque floor elevations increase northwards as downcutting fails to keep pace with uplift. Zone B includes a readily identifiable glacial influence in the form of cirques and nascent valley floors, whose upper extent occurs at generally lower elevations (Figure 2) as, in long profile, the valley floor–valley head change of slope becomes more pronounced. This influence increases northwards.

The Zone B to C transition is abrupt and is reflected in both landform and morphometry. Morphologically, the change involves the appearance of ‘classic’ glacial troughs, whose increasing radii of curvature to the north demonstrate adjustment to greater ice discharges. Cirques have been removed by erosional widening of the valleys, and wider trough floors with more pronounced long-profile concavities provide the low-altitude modal class. Ridge crests have sharpened further to become pinnacled arêtes in many places, which the constant maximum altitude suggests may lower more quickly. Morphometrically, Zone C marks the appearance of a dominant hypsometric mode at low elevations, corresponding to a marked increase in skewness and reduction in both mean and modal altitude, though not in summit elevations.

The overall morphological–morphometric pattern demonstrates a change in dominant valley-forming process, from fluvial in the south, through cirque glaciation, to full valley glaciation in the north. Although maximum relief increases throughout, trough development causes lower mean and modal elevations as valley floors enlarge to replace high plateau remnants and rounded divides as the dominant hypsometric element in the landscape.

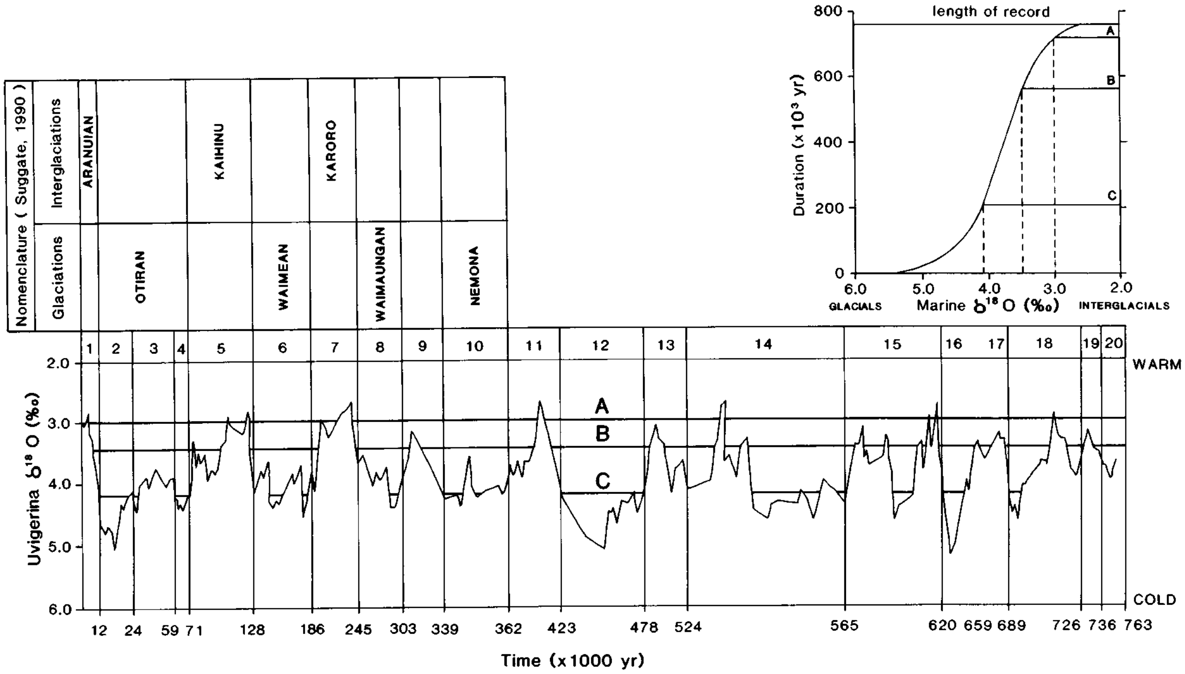

Figure 4. Variation in oxygen isotope ratio over the last 763 000 years from DSDP core 594 (for location see Figure 1). SPECMAC time-scale is from Imbrie *et al*. (1984). New Zealand nomenclature for terrestrial Glaciations and Interglaciations is from Suggate (1990). Lines A, B and C show the $\delta^{18}O$ values for the Interglacial, Intermediate and Full Glacial scenarios described in the text. Inset: curve showing the duration of time equalled or exceeded by given isotope ratios in core 594 for the entire length of record.

DURATION OF GLACIAL EROSION

The duration of glacial erosion of any valley depends on (a) the length of time that surrounding ridgelines have had sufficient altitude and proximity to the Main Divide to lie above the glaciation limit, and (b) the proportion of glacial to non-glacial climate during that time.

Climatic constraints on glaciation

To estimate glacier occupancy of the Ben Ohau Range, we combine glacio-geomorphological reconstructions by Porter (1975) with the $\delta^{18}O$ record from DSDP site 594 (Figure 4), assuming a relationship between $\delta^{18}O$ value and terrestrial ice cover. Three glacial ice cover scenarios have been selected from Porter's (1975) glacier reconstructions of the Tasman Valley. An Interglacial scenario (line A) is assumed to be typified by present-day glaciers and $\delta^{18}O$ value (3·0‰); a Full Glacial scenario (line C) by the Tekapo Advance of *c.* 12 ka BP ($\delta^{18}O$=4·2‰), and an Intermediate scenario (line B) by the early Holocene Birch Hill Advance (*c.* 8 ka BP (McSaveney and Whitehouse, 1988); ($\delta^{18}O$=3·45‰). The sedimentary record contained in a core from site 594 represents a total of nine glacial cycles in the Southern Alps during the last *c.* 0·75 Ma (Nelson *et al*., 1986; Suggate, 1990). From the inset in Figure 4, it may be seen that climate has favoured full glacial cover over *c.* 200 ka since the start of this record, while intermediate conditions prevailed or were exceeded over *c.* 550 ka.

Glacial–geomorphological reconstructions (Porter, 1975) show ice cover controlled by the intersection of range crests with a variable equilibrium line altitude (ELA) surface, tilted to the northwest at all times. The northern end of the range is thereby glaciated more often and for longer than the southern end, with a southward spread of glaciers following lowering of the ELA. During the Late Pleistocene Tekapo Advance, glaciers from Jacks Stream northwards were confluent with ice occupying the Tasman Valley. Further south, independent valley glaciers mostly reached the range foot. During the Birch Hill Advance (the Intermediate scenario), confluent glaciers occurred north of Birch Hill Stream. From Freds Stream south to Jacks Stream glacier termini

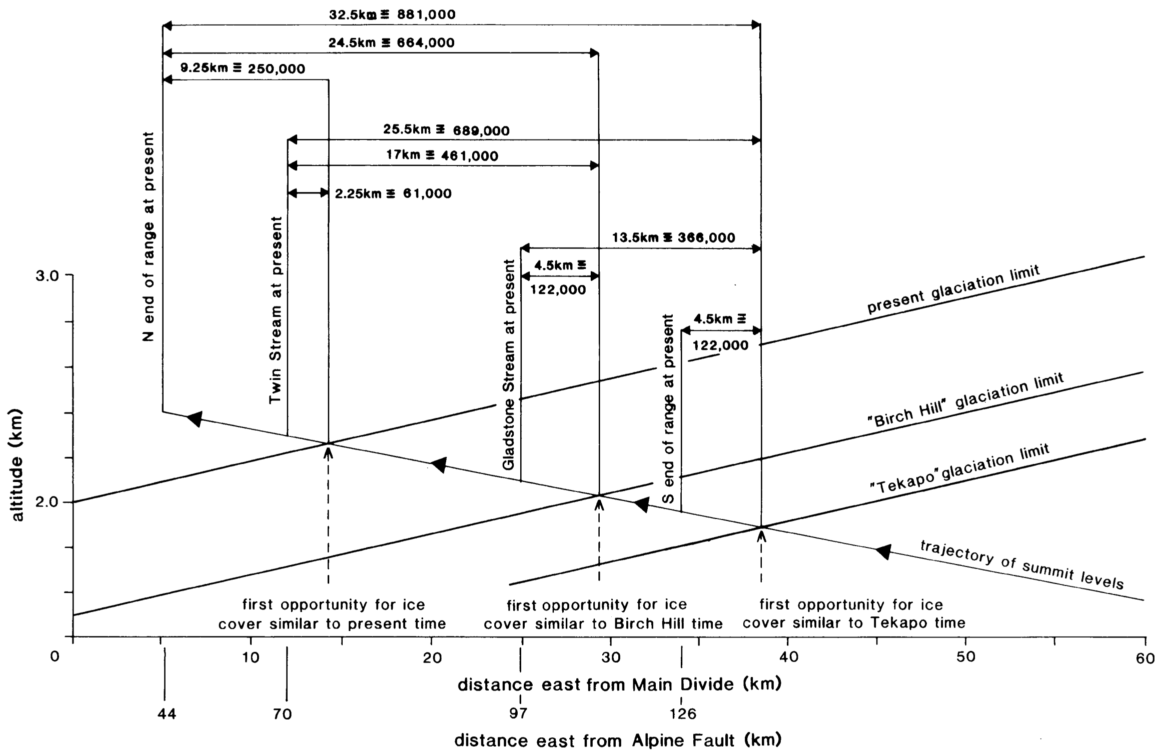

Figure 5. Projection of summit trajectory and glaciation limits onto an east–west oriented plane. Glaciation limits are plotted as (ELA-200 m), based on the empirical relationship between ELA and glaciation limit. Summit trajectories are based on tectonic transport of 39 mm a^{-1} and surface uplift of 0·6 km Ma^{-1} (Tippett and Kamp, 1995a). The diagram shows four present-day locations with respect to the long-term trajectory, and estimates of time elapsed since the earliest opportunity for glaciation under three different ice cover scenarios (after Porter, 1975).

lay mid-way down the valley floors, but south of Jacks Stream only high-level cirques and shelf glaciers formed at that time. During Interglacials, no valley floors were ice-covered.

Periods of trough cutting would therefore only have been climatically possible whenever ice cover equalled or exceeded the Intermediate scenario ($\delta^{18}O \geq 3{\cdot}45$‰) about as far south as Twin Stream. Further south, trough cutting will only have been possible during the nine phases of Full Glacial conditions ($\delta^{18}O \geq 4{\cdot}20$‰). Ice extent during all Intermediate phases may not have been replicated by ice cover in the Birch Hill period, but the southern boundary of trough cutting may have shifted with time over the central part of the range.

Tectonic constraints on glaciation

The Southern Alps form the leading edge of the Pacific Plate, moving rapidly westward and uplifting along its western margin. To estimate the tectonic transport of the Ben Ohau Range over Late Quaternary time, the plate motion vector given by de Mets *et al.* (1990) of 071° with displacement of 39 mm a^{-1} WSW has been adopted in the following calculation.

Long-term denudation can be assumed to equal rock uplift in rapidly uplifting convergent plate margins, so that fission-track ages from rocks exposed at the surface allow reconstruction of uplift history (Tippett and Kamp, 1995a,b). Tippett and Kamp found that uplift of the crust, summits and valleys appears remarkably linear over time, but that differential erosion between summits and valleys produces different surface uplift rates for each. From Tippett and Kamp's (1995a) regression equations, the trajectories followed by summits in the Ben Ohau Range have been calculated to determine the time elapsed since different parts of the range first intersected the glaciation limits of the three ice cover scenarios (Figure 5). For each glaciation limit, the point of

Table III. Estimate of duration of valley floor erosion for four locations* in the Ben Ohau Range. All values are thousands of years. First Glaciation Opportunity (FGO, row 1) is illustrated in Figure 5. Duration of climate since FGO (row 2) is estimated from Figure 4, then the actual duration of valley floor erosion (row 3) is made using ice-cover scenarios for each climate based on reconstructions by Porter (1975).

	Full Glacial climate				Intermediate climate				Interglacial climate			
	NE	TS	GS	SE	NE	TS	GS	SE	NE	TS	GS	SE
Time since FGO	881	689	366	122	664	461	122	0	250	61	0	0
Duration of climate since FGO	240	220	61	24	493	337	92	0	206	55	0	0
Estimated duration of valley-floor erosion	240	220	58	24	450	?	?	0	0	0	0	0

* NE=north end of the range; TS=Twin Stream; GS=Gladstone Stream; SE=south end of the range

intersection with the summit trajectory represents altitude and distance from the Main Divide at which glaciation first becomes possible. The rate of plate motion allows the distances of any present point from its point of 'first glaciation opportunity' (FGO) to be converted to subsequent time elapsed. For example, Figure 5 shows the present space–time location of the northern and southern ends of the range, and of Gladstone and Twin Streams, relative to their locations at the time when glaciation would first have been possible under the three ice cover scenarios. The actual duration of glaciation in each part of the range will depend on the duration of warm and cold climates since the first glaciation opportunity, which have been measured from Figure 4 (see Table III). For comparison, the distance of each of these four points east from the Alpine Fault is shown in Figure 5.

DISCUSSION: DURATION OF TROUGH CUTTING AND TOPOGRAPHIC CHANGE

Figure 6 shows the duration of ice cover equalling or exceeding the Full Glacial, Intermediate and Interglacial extents approximated by Porter's (1975) reconstructions, and the best estimate of the actual duration of glacial erosion of valley floors along the range. The figure was derived by the following stages. First, for any point on the range, the time elapsed since FGO is calculated for each of the three climate scenarios (Figure 5). Then, the actual duration, since FGO, of climates at least as cold as each climate scenario is calculated from lines A to C in Figure 3. Finally, the duration of valley-floor erosion is estimated by referring to Porter's (1975) maps of reconstructed ice cover. Thus, in the south of the range only Full Glacial conditions allow glaciers to cover valley floors, therefore the duration of trough cutting equates to the duration of Full Glacial climates since FGO. Consequently, southern valleys have only been glaciated during the last (Otiran) glaciation. Conversely, in the north, glaciers filled the valleys under Intermediate climates but not during Interglacial climates, so the duration of trough cutting equates to the duration of climate since FGO at least as severe as the Intermediate scenario. Table III summarizes calculated durations for four locations on the range. The duration of trough cutting is poorly constrained in the central part of the range, because glaciers would have filled the valleys under Full Glacial conditions but probably not during all periods of Intermediate climate. Maximum and minimum estimates are defined by these two cases (Figure 6, shaded area).

The estimates of trough cutting represent maximum durations rather than exact values. Continued rise of summits above the glaciation limit will have encouraged longer glaciers to accumulate in each successive glaciation, thereby giving an overestimate of occupancy because glacier extents have been based on reconstructions post-dating the last glacial maximum.

Several observations may be made from the morphological change from rounded, fluvially incised upland to an arête-and-trough landscape. First, the relative relief is established very early by fluvial incision, and the role of glaciers seems to be one of reshaping of valleys rather than of primary downcutting. Second, the altitude of the upper limit of gentle valley floors declines as glacial erosion becomes more marked, perhaps indicating that the rate of downcutting increases with greater glacial influence. This would accord with the northward increase

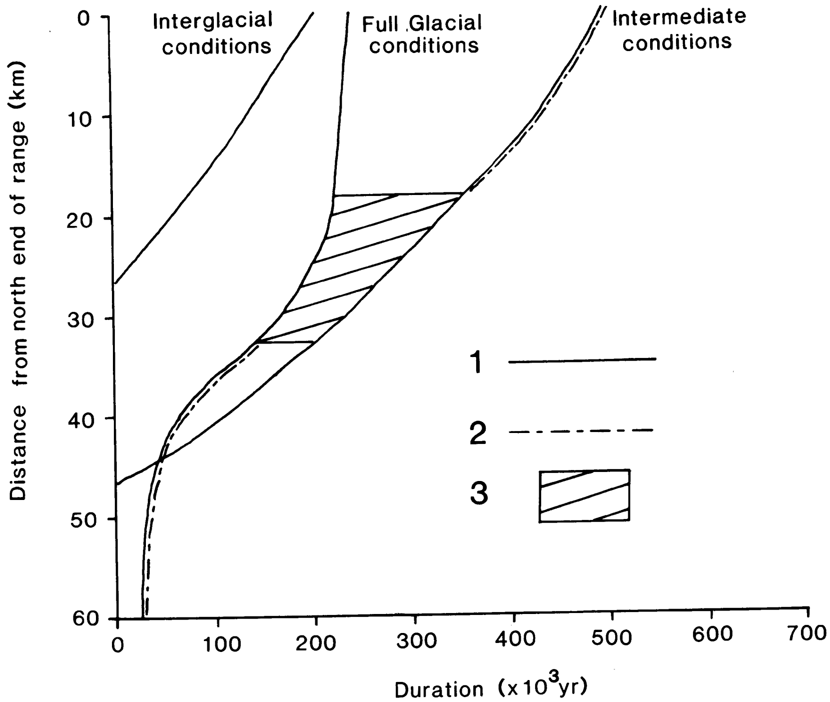

Figure 6. The variation in the duration of valley glacier cover with distance along the range, measured from the present northern extremity of the range. Estimates are best constrained in the northern and southern thirds, and less so in the central third of the range (shaded). 1=duration of ice occupancy under each climate scenario; 2=duration of glacial erosion along entire valley floor; 3=region of uncertainty of duration of trough erosion. 2 and 3 diverge in the central part of the range where glaciers did not wholly occupy the valley floor under Full Glacial and/or Intermediate scenarios.

in glacier discharge due to the climatic gradient. Third, Zones A and B suggest that cirques have evolved from sloping bench or shelf forms rather than from concave hollows.

In the Ben Ohau Range, most of the landscape of Zone A has been subjected to less than 24 ka of predominantly cirque glaciation. Valley form begins to show minor, but recognizable, glacial modification after *c.* 70 ka of occupancy, with the sharp ridgelines and pronounced cirques of Zone B becoming established after *c.* 200 ka and trough forms especially after *c.* 250 ka. Zone C follows at least *c.* 320 ka of valley glaciation. Thus, 'classic' glaciated valley landscapes cut in metasediments would appear to require at least two full glaciations to develop. This accords well with Harbor *et al.*'s (1988) order-of-magnitude estimate. The mean summit elevation of *c.* 2·4 km in the northern Ben Ohau Range indicates a total vertical denudation of 0·8 km from summits and 2·4 km from valleys, assuming a level initial surface at 4 Ma BP (Tippett and Kamp, 1995b, Figure 7). Since the earliest glaciation, vertical denudation of summits and valleys will have been 0·16 km and 0·48 km, respectively, or 20 per cent of the total. Thus, most denudation was achieved in a preglacial fluvial environment.

The strong precipitation gradient along the range will correspond to a change from glaciers with low mass-balance gradients and erosive energy to glaciers of faster throughput and high erosive energy towards the north (*cf.* Andrews, 1972). Thus, for a given period of occupancy, glaciers in the north will have performed more landscape modification than those further south. The rapid evolution of valley morphometry will result from a combination of increasing occupancy and greater erosivity of ice to the north. However, had the range never been glaciated, there would also have been a northward increase in erosion rate due to the steeper, longer slopes and higher rainfall. It is not therefore possible to compare relative glacial and fluvial erosion rates.

CONCLUSIONS

This study demonstrates the relationship between geomorphological and morphometric characteristics of a mountain range during the transition from a fluvial to a glacial landscape. Most denudation was accomplished by preglacial fluvial erosion, selectively incising an initial plateau to close to base level. Subsequent cirque development at a high level produces recognizable, but shallow, cirque forms after at least 24 ka. An increase in mean elevation contrasts with a decrease in modal elevation as the glacial influence on the uplifting range increases. The estimated time-scale for the conversion of fluvial to glacial troughs is 200 to 300 ka, confirming the numerical simulation of the transition by Harbor *et al*. (1988).

The implications for the landscape evolution model of Molnar and England (1990) are threefold. First, our hypsometric results concord qualitatively with Molnar and England's (1990) hypothesis of increasing maximum elevation coeval with decreasing mean elevation following glacial incision. Second, our results suggest that it is highly unlikely that the doubling of maximum elevation suggested by Molnar and England (1990, p. 31) could result from the isostatic response to glacial incision alone, because this would require valley cross-profiles of extreme U-shape, not possible under natural conditions of rock mass strength and slope angle (*cf*. Augustinus, 1992), and limited local relief prior to glaciation. Third, the timing of the transition to decreasing mean relief comes *after* the incision of the initial plateau to local base level. In the Ben Ohau Range, even the southernmost rivers are incised to base level, indicating fluvial incision to base level early in the period of range uplift, and destruction of the initial plateau surface prior to the onset of glaciation. There is no justification for invoking glaciation as a trigger for an isostatically induced rise in summit elevations, because incision to local base levels predates the onset of glaciation, and ridgeline elevations are unaffected by the fluvial–glacial transition in the study area.

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